Analysis of the soil moisture behaviour during wetting and drying episodes from TDR data collected at Pallanzeno (Toce Valley)

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Introduction

The MAP-Hydrology working group has different teams that carried out their experimental activities in the Lago Maggiore Target Area (Ranzi, 2000); this increased the synergy and the results of the group. In this paper some results derived from the soil moisture readings collected at Pallanzeno (Toce Valley, Italian alpine area) from April to November 1999 are reported. The soil moisture data were obtained using buriable TDR probes installed in the first 70 cm soil layer; the temporal resolution was four hours (Menziani et al., 2000). The long record of data shows some main characteristics: a daily cycle, the soil moisture trend during the driest period (July) and the soil moisture behaviour during the major precipitation event (IOP-02). Fig.1a shows the soil moisture trend at 5, 15, 25, 40 cm depths for July 1999; the bars are the four hours cumulative precipitation. In the second part of July the precipitation is practically absent up to 28 when there was a heavy precipitation. During this period, while the soil moisture at 40 cm shows a slight decrease, all the upper probes show a steep decrease crossing the 40 cm curve and reaching the lowest values of the total period of measure. In particular, on July 27, the 5 cm probe measured a soil water content of 17.7 % that was the minimum value of the whole record. Fig.1b shows the soil moisture trend at 5, 15, 25, 40 cm depths for September 1999; the bars are the four hours cumulative precipitation. In September took place the major precipitation event (MAP IOP-02). From 19 to 21 a cumulative precipitation of 225.4 mm was recorded at Pallanzeno and the highest soil water content (56.0 %) was measured at 5 cm depth on September 21. As will be shown in the following, the soil moisture behaviour after the precipitation is better described by the vertical profile trend than by the temporal trend of Fig.1b. In the first part of September the precipitation was almost absent; the soil moisture decreased at all the depths but the shallowest probes didn't reach soil moisture values lower than the 40 cm value. Finally both Fig.1a and Fig.1b show a daily cycle of the soil moisture at all the depths.

The daily oscillation

The daily oscillation, present in the experimental soil moisture data at all the levels, clearly shows a non-constant amplitude versus the soil depth while the time delay results quite independent from the depth. To explain this behaviour a laboratory experiment has been carried out; both a sandy soil at saturation and the instrument were brought to temperatures significantly higher and lower (about ten degrees) respect to the ambient one. This experiment allowed verifying the TDR soil moisture measurement dependence on both soil temperature (Pepin et al., 1995) and instrument temperature. For the considered experiment a linear relationship between the soil moisture change and the temperature change (of the soil and of the instrument) has been found. In particular the slopes of the regression lines resulted -0.08 and -0.10 (%/°C) for the sandy soil and the instrument respectively. At Pallanzeno site the soil temperature profile was not measured. Using the air temperature as the temperature of the instrument (most likely underestimated) the daily oscillation is well reproduced by means of coefficients depending on the soil moisture. This implies that the daily oscillation seen in the experimental data is due to the soil conditions and to the thermal trend of the instrument.

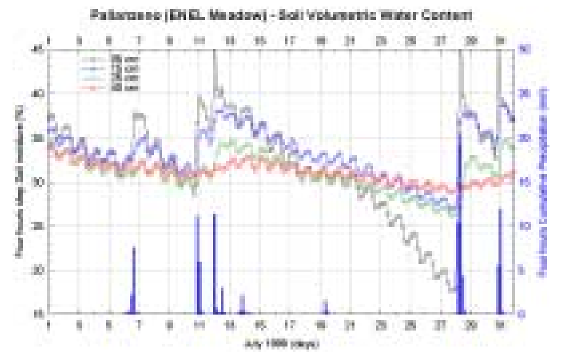

Fig.1a

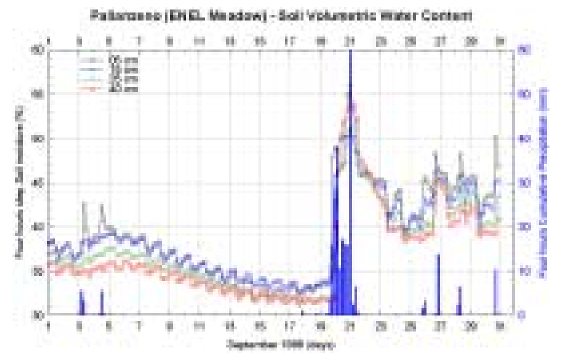

Fig.1b

Soil physical characteristics

The soil physical properties and pedologic characteristics of the Pallanzeno site were obtained from in situ and in laboratory measurements performed by the University of Brescia (Falappi et al., 2000) and Istituto Agrario di San Michele all'Adige (ISMA) (Eccel et al., 2000).
The soil moisture characteristic and the unsaturated hydraulic conductivity can be described respectively by the following power functions (Campbell, 1985)

$$\psi = \psi_s \left(\frac{\theta}{\theta_s} \right)^{-b} \quad ; \quad k = k_s \left(\frac{\theta}{\theta_s} \right)^{2b+3} \qquad (1).$$

Where Ψ_s is the air entry water potential; θ_s is the saturation water content; k_s is the saturated hydraulic conductivity of the soil. Ψ_s and b are the offset and the slope of the log-log retentivity curve. From the soil sample collected at Pallanzeno three retentivity curves for three different layers (0-10, 10-20, 20-40 cm) were obtained. The values of Ψ_s and b for the three previous quoted layers are reported in Tab.1.

Table 1

Depth (cm)	Air entry potential (m)	b
0-10	0.33	4.2
10-20	0.14	4.3
20-40	0.07	4.6

The hydraulic conductivity at the saturation (k_s) was measured in laboratory (sample of the first 0-14 cm layer) and in situ by means of an infiltration test representative of a deeper layer. The real value of k_s is expected to be between the measured values reported in Tab.2 and, due to the difficulties related to the in situ test, most likely closer to the laboratory value (Falappi et al., 2000). The soil water content at the saturation (θ_s) has been assumed to be 7.5 % less than porosity; this value equals the maximum soil water content measured at Pallanzeno during the IOP-02 (see Tab.2).

Table 2

Parameter	Value
k_s (laboratory)	$2.26\ 10^{-7}$ (m/s)
k_s (in situ)	$1.89\ 10^{-4}$ (m/s)
ϕ (porosity)	0.61
$\theta_s=0.925\ \phi$	0.562
θ_{max} (measured)	56 (%)

Soil moisture trend during the driest period

The soil water content is a function of space and time; considering only the vertical direction it can be obtained solving the one dimensional Richards equation. Introducing $\theta' = \theta/\theta_s$ (ranging from 0 to 1) and defining the hydraulic diffusivity as $D = k \cdot \partial\psi/\partial\theta'$ the Richards equation can be written as

$$\frac{\partial\theta'}{\partial t} = \frac{\partial}{\partial z}\left(D \cdot \frac{\partial\theta'}{\partial z} \right) - \frac{\partial k}{\partial z} \qquad (2)$$

Assuming valid equations (1) and considering only one layer (from the surface up to 40 cm) having the mean soil physical characteristics derived from Tab.1 and a mean k_s of $3\ 10^{-6}$ (m/s), equation (2) has been numerically solved. The initial condition (uniform) and the upper and lower boundary conditions were obtained from the experimental data. Fig.2a shows the results of the numerical solution of equation (2); the evolution of the vertical profile of the soil moisture from 19 to 27 July 1999 (driest period) is compared with the experimental data (circles). The vertical profiles have been extrapolated assuming a linear trend from 5 cm to the soil surface.

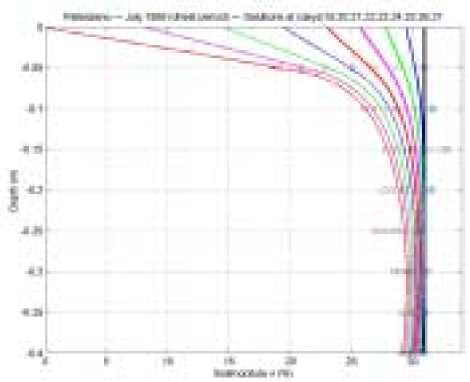

Fig.2a

Fig.2b shows the water loss from a column of unitary cross section and 40 cm height; the solid line is the cumulative loss computed from the numerical solution and the circles represent the one obtained from the experimental data. Under the assumption that the loss of water is essentially from the soil to the atmosphere, it represents the cumulative evaporation.

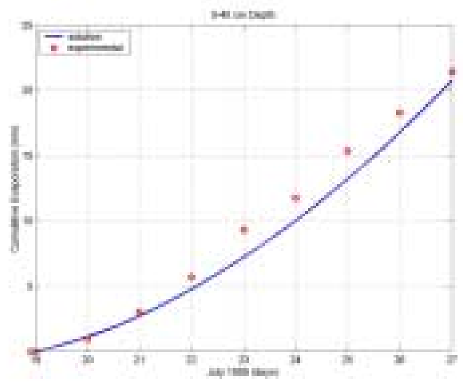

Fig.2b

One of the main problems to analytically solve the Richards equation is that the hydraulic diffusivity depends on the soil water content itself. Usually, when the evaporation is the most important process concerned, the $\partial k/\partial z$ term in equation (2) can be neglected and equation (2) becomes the diffusion equation. The diffusion equation can be easily analytically solved considering a constant hydraulic diffusivity and the desorptive diffusion approximation (Brutsaert and Chen, 1995). The boundary conditions for desorption implies that θ(z=0,t>0) equals the final surface wetness while this is not the experimental case (see Fig.2a). In this paper the analytical solution of the following equation (3) is presented.

$$\frac{\partial \theta}{\partial t} = \overline{D} \cdot \frac{\partial^2 \theta}{\partial z^2} - k_v \frac{\partial \theta}{\partial z} \qquad (3)$$

The mean diffusivity to be used has to characterise the drying process. According to Crank (1956), the weighted mean diffusivity for desorption processes is

$$\overline{D} = \frac{1.85}{\left(\theta_i - \theta_f\right)^{1.85}} \cdot \int_{\theta_f}^{\theta_i} D(\theta) \cdot \left(\theta_i - \theta\right)^{0.85} d\theta \qquad (4)$$

Considering for the soil water content the following initial and final values (θ_i=30 % and θ_f=10 %) and taking into account that 4<b<5, the mean diffusivity ranges from 4.5 to 8.4 10^{-9} (m^2/s). The second term on the right of equation (3) is a negative source term proportional to the vertical gradient of the soil moisture. It can be considered as the contribution of the vapour phase. The constant k_v has the dimension of a velocity. The following equation (5) is the

$$\theta = \theta_i - \frac{\theta_i}{2} \cdot erfc\, \frac{z - k_v \cdot t}{\sqrt{4 \cdot \overline{D} \cdot t}} - \frac{\theta_i}{2} \cdot e^{k_v \cdot z/\overline{D}} \cdot erfc\, \frac{z + k_v \cdot t}{\sqrt{4 \cdot \overline{D} \cdot t}} \qquad (5)$$

analytical solution of equation (3) which implies a uniform initial condition and the boundary condition at z=0 described by equation (6)

$$\theta(z=0,t) = \theta_i \cdot erfc \sqrt{k_v^2 \cdot t / 4 \cdot D} \qquad (6)$$

Fig.3a shows the evolution (19 to 27 July 1999) of the vertical profile of the soil moisture obtained from the analytical solution (equation (5)). The solution shown in Fig.3a has been obtained using D=5.9 10^{-9} (m^2/s) and $k_v=10^{-7}$ (m/s). The dots are the experimental data. The agreement between the theoretical curves and the experimental data is good mainly in the upper part of the soil layer. Fig.3b shows the water loss from a column of unitary cross section and 15 cm height; the solid line is the cumulative loss computed from the analytical solution and the circles represent the one obtained from the experimental data.

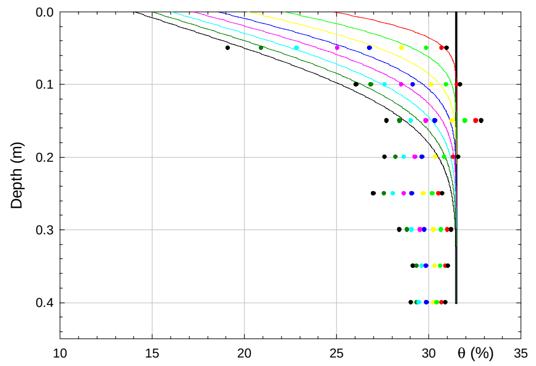

Fig.3a

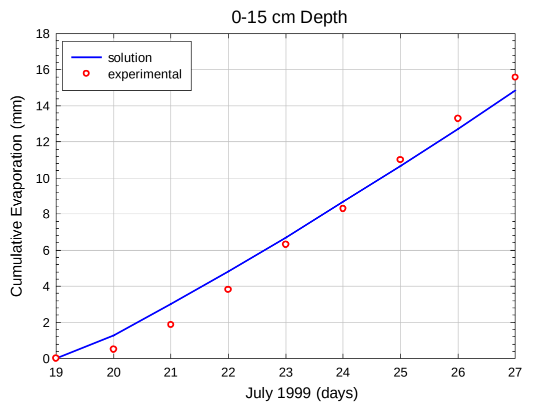

Fig.3b

Soil moisture trend during the IOP-02

Using the differential equation (3), substituting V to k_v and assuming as initial condition $\theta=\theta_i$ and as boundary condition $\theta(z=0,t>0)=\theta_f$, a solution comparable with the experimental soil moisture profiles of the wet case IOP-02 is found

$$\theta = \theta_f - \frac{\theta_f - \theta_i}{2} \cdot erfc\, \frac{V \cdot t - z}{\sqrt{4 \cdot \overline{D} \cdot t}} + \frac{\theta_f - \theta_i}{2} \cdot e^{V \cdot z/\overline{D}} \cdot erfc\, \frac{V \cdot t + z}{\sqrt{4 \cdot \overline{D} \cdot t}} \qquad (7)$$

The mean diffusivity has been computed using a relationship similar to equation (4) but suitable for the infiltration process (Crank, 1956).

The V constant of equation (7) has the dimension of a velocity. The second derivative of the function expressed by equation (7) equals zero in the range

$$\boxed{V \cdot t < z < 2 \cdot V \cdot t}$$

This implies the presence of an inflection point, which moves downward with a velocity depending on V. The initial soil moisture value (33 %) corresponds to the mean value of the first experimental profile (September 19 at 16:00 LST). From the experimental data the final soil moisture value has been chosen 49 %. In Fig.4a the experimental profiles (see the legend) at four hours intervals are compared with the theoretical results (solid lines) obtained from equation (7). The first experimental profile (September 19 at 16:00 LST) is the last collected just before the precipitation. Fig.4a shows the evolution of the soil moisture in the first 40 cm layer from September 19, 4 pm to September 20, 8 am. The water precipitated during this period (92.7 mm) produced the major change in soil moisture; that is the soil reached an almost complete saturation. The water precipitated in the second part of the event produced only a minor increase of the soil moisture. The agreement between the theoretical and experimental data is good in the first 40 cm but not so satisfactory in the lower part. On the other hand it has to be remarked that the initial soil moisture is almost uniform only in the first 40 cm layer.

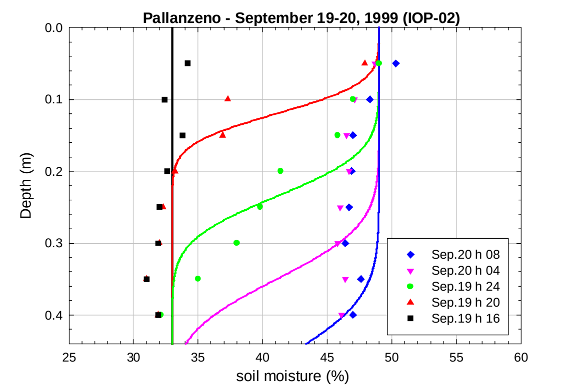

Fig.4a

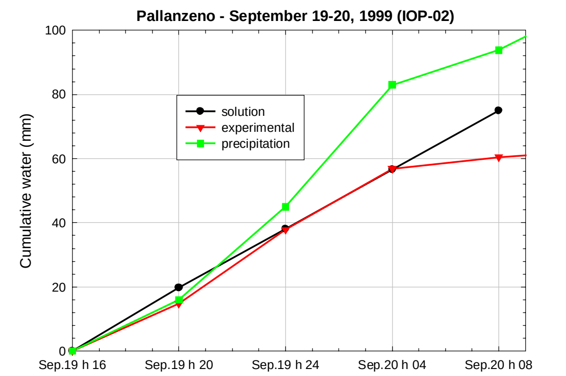

Fig.4b

Fig.4b shows the water accumulated in a column of unitary cross section and 40 cm height. The three curves represent the cumulative water computed from the analytical solution (equation (7)) (circle), from the experimental data (triangle) and the precipitated water (square). From Fig.4b it is evident that the water precipitated during these sixteen hours is greater than the water accumulated in the first 40 cm of soil.

Acknowledgements

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